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Stress transfer and its implication for earthquake hazard on the Kunlun Fault, Tibet

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Abstract

The 1600-km-long Kunlun Fault striking E-W to WNW-ESE had long been recognized as one of the major left-lateral strike-slip faults bounding the Tibetan Plateau, and ranked one of the most active faults in China continent. During the past hundred years, over twenty strong earthquakes occurred along and near the Kunlun Fault, including six large earthquakes (M>7). Since some major highly-populated and industrialized cities are close to the Kunlun Fault, understanding of stress transfer and earthquakes migration along the Kunlun Fault is most important for assessing seismic hazard in this region. In this study, by integrating coseismic effect, viscoelastic relaxation and tectonic loading, we studied the evolution of the regional Coulomb stress field by analyzing a sequence of strong earthquakes along the Kunlun Fault. We studied the stress evolution over one century by analysing a sequence of five earthquakes (M≥7) that occurred along the Kunlun Fault since 1937. The model of dislocation sources embedded in a mixed elastic/inelastic layered half-space was used, and the layered model and relevant parameters were constrained by seismic studies. Fault rupture locations and geometry, as well as slip distribution of earthquakes were taken from field observations and seismic studies.

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Numerical results showed a good correlation between stress transfer, accumulation and earthquakes occurrence. All four studied earthquakes occurred after the 1937 Tuosuo Lake quake were encouraged by the preceding earthquakes with positive stress loading. In subject to the choice of the earthquake source parameter two or three out of four events occurred in regions that experienced previous coseismic and postseismic stress changes of at least 0.01 MPa, suggesting that earthquake triggering due to stress transfer has occurred along the Kunlun Fault. The total stress change since 1937 of the Kunlun Fault region has lead to high levels of stress accumulated on the Xidatan-Dongdatan segment and Maqin-Maqu segment, which have not experienced any significant large earthquake over at least several hundred or several thousand years.. The accumulated stress raises the potential earthquake hazard in these areas. Our study demonstrated the crucial importance of postseismic viscoelastic relaxation in the stress transfer and accumulation following large earthquakes.

Keywords: Stress transfer; Stress accumulation; Earthquake triggering; Earthquake hazard

1. Introduction

The Kunlun Fault (KF), extending about 1600 km between 86°E and 105°E, is one of the largest strike-slip faults in the northern Tibet (Fig. 1) (Van Der Woerd et al., 2000, 2002a). Fieldwork confirms that the E-W to WNW-ESE striking KF is a major strike-slip fault that accommodates for both, the northeastward shortening and the eastward extrusion of Tibet (e.g., Tapponnier and Molnar, 1997; Yin and Harrison, 2000). It is suggested that the differential motion of 10-20 mm/yr between the north Tibet and south-central Tibet is mostly accommodated by the KF as seismic slip (Lin et al., 2006). As a result, the KF has experienced strong earthquakes, including six M≥7 events in the last 100 years (Fig. 1). Since there are several major highly-populated and industrialized cities close to the KF, such as Golmud and Delinhar, and recent events appear to be propagating towards some populated areas, understanding of earthquakes migration along the KF is most important for assessing seismic hazard in this region.

The earthquake sequence along the KF (Table 1) shows evidence for time-space progression, suggesting certain interaction among earthquakes. Interaction between earthquakes is suggested to realize in a manner of earthquake triggering by the change of Coulomb Failure Stress (ΔCFS) (Stein 2003): positive ΔCFS brings the fault closer to failure and thus earthquake occurrence, while negative ΔCFS retards subsequent events (Stein 1999; Freed, 2005).

A number of studies have successfully used the Coulomb model to explain aftershock distribution (King et al., 1994; Reasenberg and Simpson 1992; Parsons et al. 1999; Toda et al., 1998; Wyss and Wiemer 2000; Ma et al., 2005), earthquake sequences (Stein et al. 1994; Hodgkinson et al. 1996; Nalbant et al. 1998), and triggering of moderate to large earthquakes (Harris et al., 1995; Deng and Sykes, 1996; Jaume and Sykes, 1996; Martínez-Díaz et al., 2006) as well as to assessing earthquake risk (McCloskey et al., 2005; Nalbant et al., 2005). However, a limitation of most of these models is that they only consider elastic responses to fault slip, and thus cannot account for delay times in the triggering process (Freed and Lin, 2001; Freed 2005). Postseismic stress changes due to viscoelastic stress relaxation in the lower crust and/or upper mantle were taken into account to explain aftershock distribution and the triggering of later events at the time scale of years and decades (Deng et al., 1999; Freed and Lin, 2001; Pollitz and Sacks, 2002). Recently, a number of models incorporating postseismic viscoelastic relaxation have been developed and applied successfully on study the triggering of earthquake pairs and sequences (Pollitz et al., 2003; Lorenzo-Martín et al., 2006; Ali et al., 2008)

Similar elastic (Chen et al., 2003) or viscoelastic Coulomb models (Shen et al., 2003) were published for the KF region. However, some source parameters and the rheological model commonly employed by these works were not well constrained. New evidence from further studies carried out in the last few years should be incorporated to constrain the co-seismic slip, rock properties and stratification. Moreover, various mechanisms responsible for ΔCFS should be incorporated toward a comprehensive understanding of the earthquake interaction. Therefore, a study integrating coseismic and postseismic stress change together with tectonic loading (e.g., Lorenzo-Martín et al., 2006; Freed et al., 2007; Ali et al., 2008) is expected to re-evaluate the process of stress evolution and nature of earthquake on the KF.

In this study, we investigated a sequence of five earthquakes of M≥7.0 since 1937 (Table 1 and fig.1) on the KF. We studied the stress evolution since 1937 by integrating coseismic slip, postseismic viscoelastic relaxation of the lower crust and mantle, and interseismic tectonic loading due to India-Eurasia convergence. In particular, we evaluated the importance of the individual process on the total stress field. Instead of fixing the average strike direction of the KF, we took into account the details in orientation of the different ruptures and different segments of the fault. The evolution of the ΔCFS at the hypocenters of the shocks, as well as the state on the rupture surfaces immediately before the earthquakes were examined under different assumptions of the rheology. In addition, we extrapolated the study to 2040 to evaluate the stress state on various segments of the KF. The aim of the study is to investigate the relationship between ΔCFS and seismicity, with emphasis on identifying segments on the KF that have experienced large build-up of unrelieved stress to provide useful information for seismic hazard assessment in this region.

2 Earthquake sequence and source parameters

Left-lateral motion along the KF is believed to take up a substantial fraction of the northward motion of the Indian plate under Tibet by left-lateral strike-slip. This motion probably initiated in the late Miocene or early Pleistocene (Kidd and Molnar, 1988; Fu and Awata, 2004). The total slip along the KF is about 75 km, based on the offset of a meta-sedimentary unit interpreted from Landsat images (Kidd and Molnar 1988). The late Quaternary slip rate on the KF is 12±3 mm/yr, derived from cosmogenic dating of offset stream risers. This estimation is supported by further cosmogenic surface dating and radiocarbon dating (Van der Woerd et al., 2002b; Li et al., 2005), and trenching surveys (Zhao, 1996), as well as GPS measurements (Wang et al., 2001). Given a uniform rate of 12 mm/yr since its initiation, the magnitude of slip along the KF implies that the KF has been active over the past 7 Ma (Yin and Harrison, 2000).

However, no earthquake was instrumentally recorded on the KF before the 1900s. During the twentieth century, the seismicity of the KF was sparse but continuous with several moderate and large earthquakes. There are 19 earthquakes of M≥5 occurred on the Kunlun fault since 1900: 6 events of M≥7, 1 of M=6.3 and 12 of M<5.5. Since it is suggested that moderate earthquakes (6.0≤Mw<6.5) only perturb the stress locally (10s of km) (Freed et al., 2007) and given that our focus is on the evolution of stress over a broad region, it is reasonable to assume that local perturbation of small to moderate earthquakes are insignificant for the overall stress pattern. Therefore, in the present study, we only considered the earthquakes of M≥7 in our analysis.

At the beginning of the last century, a strong earthquake of M7.0 shook Xiugou on November 4, 1902 (SBQP, 1999), but few information of this earthquake is available, except the poorly located epicenter. We have to exclude it from the earthquake sequence for analysis. The subsequent, the 1937 Tuosuo Lake (Huashixia) earthquake (M7.5) was particular large. It ruptured the Tuosuo Lake (or Dongxi Co) segment of the fault, producing a surface rupture with a length of 150-240 km long and a left-lateral slip ranging 4-7 m (Liu, 1999; SBQP, 1999; Van der Woerd, et al., 2002b; Guo et al., 2007). Since the influence of the earthquakes before 1937 is difficult to be incorporated in the analysis due to the lack of information, we set the 1937 earthquake as the first of the studied earthquake sequence, and confine our purpose and interest in the interaction among the 1937 earthquake and subsequent events to study the interaction among them.

Twenty-six years later, the M7.0 Alake Lake earthquake occurred on the Alake Lake segment west of the 1937 coseismic rupture segment (SBQP, 1999), causing ~40 km-long rupture with 1-2 m left-lateral slip (Guo et al., 2007). The second largest earthquake in KF region is the 1997 Mw7.6 Manyi earthquake, which was proceeded by the 1973 M7.3 Manyi earthquake occurred along the western part of the Manyi fault, a branch fault of the

Kunlun fault system (Molanr and Chen, 1983; Velasco et al., 2000). The 1997 Manyi earthquake produced a 170-km-long surface-rupture zone along the westernmost strand of the KF (~86°E to 88°E) with a maximum slip of 7 m (Peltzer et al., 1999; Van der Woerd et al., 2002a,b; Xu 2000). It ruptured a fault which we interpret to be one splay of the KF horse-tail west of 91°E (Van der Woerd, et al., 2002b). The last and largest earthquake is the 2001 Mw7.8 Kokoxili (Kunlun) earthquake that produced a 400 km-long surface-rupture zone along the Kusai lake segment (Lin et al., 2002, 2003; Van der Woerd, 2002a; Xu et al., 2002; Fu and Lin, 2003; Fu, et al., 2005; Lasserre et al., 2005) between the 1937 Tuosuo lake and 1997 Manyi surface rupture zones. The rupture length of Kokoxili earthquake ranks the longest coseismic surface rupture for an intracontinental earthquake ever recorded.

The earthquake sequence of these five events provides an excellent chance to decipher how ΔCFS on the KF evolved over the past several decades and how earthquakes communicated with each other by stress transfer. On the other hand, it also provides a good opportunity to outline the segments which have experienced large build-up of stress.

Source information of the earthquakes, such as focal mechanism and slip distribution, is of crucial importance for stress analysis and therefore must be examined carefully. For 1937 Tuosuo Lake earthquake, the length and slip distribution of the rupture is debated. While some authors suggested a longer rupture zone (~240-300 km) (Li et al., 2006; Molnar and Deng, 1984), others favoured a shorter dimension of the rupture (150-180 km) (Guo et al., 2007; Van der Woerd et al., 2002a). Both, homogeneous co-seismic slip (Guo et al., 2007) and distributed co-seismic slip (Li et al., 2006) models were proposed. Since convincing evidences are not available due to degradation and overlap of older earthquakes, it is difficult to discern a more reliable one from several candidate models. In this study, we calculated the stress changes by using the models proposed by Guo et al

(2007) and Li et al (2006), and analysed the difference between the results. Strike, dip and rake were chosen based on the focal mechanism determined by Molnar and Deng (1984).

The length and slip of the rupture segment of the 1963 Alake Lake earthquake were taken from Guo et al. (2007). We inferred width of rupture by estimating rupture areas with the empirical scaling laws and relationships of Wells and Coppersmith (1994). The focal mechanism was taken from Molnar and Lyon-Caen (1989).

For 1973 Manyi earthquake, since the length, width and slip are poorly constrained, we have to look for an alternative way. We modelled the earthquake as rectangular planar patches with uniform slip occurring within an elastic crust. The thickness of the elastic crust was assumed to be the same as that determined in the 1997 Manyi earthquake. Then, we used the magnitudes together with the empirical scaling laws by Wells and Coppersmith (1994) to estimate the rupture area and the slip amplitude. Strike, dip and rake were taken from Molnar and Chen (1983).

The parameters of the Manyi earthquake determined by different authors differ significantly from each other. For example, the rupture length has been estimated to be 47 km (Liu et al., 2000), 70 km (Xu and Chen, 1999), 110 km (Liu et al., 2002) and 170 km (Peltzer et al., 1999; Funning et al., 2007), respectively. The width ranges from 18 km (Funning et al., 2007), 28 km (Liu et al., 2000) to 63 km (Xu and Chen, 1999). Among the various source models, the rupture length given by Liu et al (2000) and Xu & Chen (1999) (47 km and 70 km, respectively) is much shorter than what was obtained by field survey (Xu, 2000). And inconsistently, the estimated seismic moment by these authors is greater than all other studies (e.g., Funning et al., 2007; Shan et al., 2002; Wang et al., 2007; Peltzer et al., 1999). Therefore, the source models by Liu et al (2000) and Xu & Chen (1999) were excluded from the present study. The other source models (Funning et al., 2007; Shan et al., 2002; Wang et al., 2007; Peltzer et al., 1999) predicted similar rupture extents despite some slight differences of slip patterns, and more consistent surface rupture with the evidence from field study (Xu, 2000), In our stress calculation, we used two most recent models proposed by Funning et al. (2007) and Wang et al. (2007).

The rupture process and slip distribution of Kokoxili earthquake has been studied comprehensively. Although consensus has been reach concerning the rupture length, complexity of slip partitioning, detailed slip distribution differs among individual studies on field investigation (e.g., Fu et al., 2005; Xu et al., 2006), InSAR imaging (Lasserre et al., 2005) and teleseismic inversion. In this study, we used 5 km×5 km gridded fault model and strike-slip distribution proposed by Lasserre et al. (2005) to calculate the stress field. Other slip distribution models (Fu et al., 2005) were used to test the stability of numerical results. The source parameters associated with all earthquakes are summarized Table 1.

3. Model and methods

We conducted our study on the basis of the change of Coulomb Failure Stress (ΔCFS) (Scholz, 1990) using the expression

$$\Delta CFS = \Delta\tau - \mu'\Delta\sigma_N \qquad (1)$$

where τ is the shear stress, σ_N is the normal stress and μ' is the apparent coefficient of friction. The change in shear stress $\Delta\tau$ is positive in direction of the slip of the following earthquake (the observing fault); $\Delta\sigma_N$ is positive for increasing clamping normal stress with pressure defined positive. The equation implies that regional faults that lie in areas of positive ΔCFS are brought closer to failure, whereas faults that lie in areas of negative ΔCFS are brought further away (Freed 2005).

In this study, we calculated the evolution of ΔCFS in the KF region by considering contributions from coseismic, postseismic and tectonic loading since the 1937 Tuosuo Lake earthquake. To calculate the coseismic and postseismic stress, we used the model of dislocation sources embedded in a mixed elastic/inelastic layered half-space (Wang et al., 2003, 2006). In contrast to Okada (1992), our model allows us to implement a more realistic rheology, so that postseismic effects can be studied. In fact, the Okada model (1992) is the special case of a homogeneous elastic half-space (see Wang et al. 2003 for details of comparison). We also employed the code PSGRN/PSCMP (Wang et al., 2006), by which surface and subsurface deformation due to the common geophysical sources in a multi-layered viscoelastic-gravitational half-space can be easily determined. In our modelling, the earth surface was treated as plane. The geometric deviation of the spherical surface from the corresponding planar surface for our studied area, that can be measured by the ratio between the arc height (~60 km) and the arc length (~ 1700 km), is about 3-4%. For such a small deviation, we may expect that its influence on the deformation field is similarly small. Consequently, the difference found so far between the spherical and planar earth models should not be dominated by the curvature effect but by the layering effect that has been considered in our analysis. Therefore, despite large extent of our studied area (~1700 km), the earth's curvature was not taken into account, although it was emphasized by some studies (e.g., Pollitz 1997).

The magnitude and pattern of postseismic deformation and stress changes depend strongly on the rheological layering of the crust and upper mantle, which in turn depends on composition and ambient temperature and pressure. Seismic data show a Moho depth of ~65 km in the studied area (Li et al., 2004). Since most earthquakes on the KF occurred at depths shallower than 20 km, and all slip models of the Manyi (e.g., Funning et al., 2007; Wang et al., 2007) and Kokoxili (Lasserre et al, 2005) earthquakes suggest that the ruptures extend down to a depth of 20 km. We therefore set the thickness of elastic crust and the locking depth to be 20 km. We assumed that viscoelastic processes occur below the depth of 20 km. Below the depth, coseismic stress changes within the viscous lower crust and upper mantle cannot be sustained and lead to visco-elastic flow, which induces stress changes in the seismogenic crust (Hergert and Heidbach, 2007; Ali et al., 2008). In our study we use the linear Maxwell rheology to study the visco-elastic effects. Although it is suggested that some other rheological models, such as SLS (Cohen, 1982; Pollitz et al., 1998; Ryder et al., 2007), power-law, non-linear rheologies (Pollitz et al., 2001; Freed and Bürgmann, 2004) were suggested could be better to analyse post-seismic relaxation processes, we, in our opinion, are not able to discern among different rheologies. Therefore, we prefer to use the simplest and most used linear Maxwell rheology, instead of more complicated models that would add additional unknowns to our analysis.

The layered model used for our calculations is described by the parameters summarized in Fig. 2. The thickness of crustal layers, density distribution, and V_P were taken from seismic studies, both tomography models (Li et al., 2006; Zhou et al., 2006) and deep seismic sounding experiments (Wang and Qian, 2000; Li et al., 2004). The quantities of density ρ and V_P were used to derive the shear modulus μ using the following expression (Aki & Richards, 2002)

$$\mu = \rho V_s^2 \approx \frac{1}{3} \rho V_P^2 \quad . \qquad (2)$$

We determined the model viscosity of lower crust and upper mantle by evidence from the studies on postseismic displacement (Ryder et al., 2007; Shao, et al., 2008). Other viscosity values were used to test the stability of numerical results.

We modeled the tectonic stress loading following a procedure outlined by Lorenzo-Martín et al. (2006) and Heidbach & Ben-Avraham (2007). The tectonic stress loading was realized by a steady slip over the depth ranging 20 to 100 km, and the deep dislocation technique proposed by Savage (1983). The fault is assumed to be locked at 20 km depth. The slip rates increase from zero at 20 km depth to its full magnitude at 68 km depth. From 68 km to 100 km depth the full slip rates were applied. The magnitude of the slip on the KF was taken from GPS interpretations (Chen et al., 2000; Wang et al., 2001) and was indicated by numbers in rectangle overlapping on segments of fault (Fig. 1). At the end segments of the fault, the tapered slip was used to minimise edge effects.

We considered a Poisson's ratio of 0.25 and set the apparent coefficient of friction μ' to a moderate value of μ'=0.4 (King et al., 1994). Different values for μ' were also tried to test the stability of results (see 4.3 for details). We present the results of the stress change calculations in terms of ΔCFS values on a horizontal plane at 10 km depth that consists of 141×221 grid points, corresponding to 3′×6′ grid spacing, covering the study region confined by 32°N-39°N and 83°E-105°E. At each stage, we calculated the ΔCFS for a the fault plane orientation of the next inspected event, instead of using optimally oriented fault planes or fixing the average strike direction of the KF.

4. Numerical results

4.1 Stress transfer and accumulation on the KF

We calculated the cumulative ΔCFS using the parameters described above and shown in the Fig. 2. The viscosities of lower and upper mantle were set to 1×10^{18} and 1×10^{20} Pa·s respectively, which were given by studies on postseismic deformation (Ryder et al., 2007; Shao et al., 2008). Slip models of Guo et al. (2007) and Funning et al. (2007) were employed for the 1937 Tuosuo Lake and the 1973 Manyi earthquake. Different source models published by other authors were used for testing the stability of the results (Section 4.2).

The evolution of the calculated ΔCFS on the KF is illustrated in the Fig.3, from the beginning of the earthquake sequence to the present. Since the hypocenters of most earthquakes in the KF region were confined within the depths of 10-20 km, and the maximum slip of the two large earthquakes, i.e. the 1997 Manyi and 2001 Kokoxili earthquakes, were commonly located at depths less than 10 km, we calculated the ΔCFS values on a horizontal plane at that depth. The maximum and average ΔCFS at the rupture surfaces are summarized in the Table 1 and the stress evolution at the hypocenters are shown in the Fig. 4.

The snapshot series begins with the first event, the 1937 Tuosuo Lake earthquake (Fig.3a). Then, the state of the stress field is presented for time immediately before each subsequent event. The 1937 earthquake that ruptured a large portion of the central KF loaded the entire rupture surface of the 1963 Alake Lake earthquake with coseismic stress over 0.01MPa (Fig. 3a). The ΔCFS immediately before the 1963 Alake Lake event is shown in Fig. 3b. The joint effect of elastic, and viscoelastic loading led to a stress increase of up to 0.06 MPa in average and 0.11 MPa in maximum on the rupture surface (Table 2), which is higher than the proposed threshold value (0.01 MPa) suggested for earthquake triggering (e.g., King et al., 1994; Stein 1999; Heidbach and Ben-Avraham, 2007). The 1963 Alake Lake earthquake was obviously encouraged by the ΔCFS associated with the 1937 event.

The coseismic stress change associated with the 1963 Alake Lake earthquake posed little impact on the overall stress pattern, except in its near-field areas (Fig. 3c). Since the 1973 Manyi earthquake was located far from the 1963 earthquake rupture, and the time period between events was too short for postseismic stress relaxation, our results show a negligible interaction between these two events (<0.005 MPa). Therefore, we suggested that the 1973 Manyi earthquake might be a result of stress accumulation due to tectonic loading.

The 1973 Manyi earthquake significantly changed the stress pattern of the western segment of the Manyi fault (Fig. 3d). The 1997 Manyi earthquake occurred directly adjacent to the 1973 Manyi earthquake. The ΔCFS increase on the rupture surface of the 1997 Manyi event is as much as 0.063 MPa in average due to the coseismic and postseismic stress of the 1973 event (Fig. 3d and Table 2).

The 1997 Manyi earthquake produced negligible coseismic stress change on the rupture surface (0.001 MPa) at the hypocenter of the 2001 Kokoxili earthquake (Table 2). However, the postseismic stress change caused by all preceding earthquakes reached 0.01 MPa at the hypocenter given by Harvard CMT, and 0.046 MPa at that by USGS. Although the stress loading was not significant, the entire rupture surface was loaded by positive ΔCFS of 0.014 MPa in average and 0.04 MPa in maximum. Particular, over 40 percentage of the entire rupture surface was stressed over 0.01 MPa (Fig. 3e and Table 2). Therefore, the Kokoxili earthquake could be encouraged by the joint effects of tectonic, elastic and viscoelastic loading produced by the preceding earthquakes.

In order to analyse the stress transfer and to examine the contributions from co- and post-seismic components in detail, we plotted the evolution of ΔCFS at hypocenters of each earthquake in Fig.4. Except for the 1937 Tuosuo Lake earthquake the 1963, 1973 and 1997 earthquakes caused coseismic stress changes < 0.01 MPa. Thus, we can infer that the static stress transfer is not the major control for the occurrence of subsequent events. In contrast to coseismic stress change, the postseismic relaxation effect plays an important role for the stress transfer, accumulation and earthquake triggering. For example, at the hypocenter of the 1963 earthquake, the postseismic stress change associated with the 1937 earthquake is five times larger compared to coseismic one (Fig. 4 and Table 2). The same holds on for the earthquakes of 1973, 1997 and 2001, in which the effects of postseismic relaxation are comparable or even much greater in magnitude than the ΔCFS caused by the tectonic loading (Table 2 and Fig. 4), suggesting a dominant role of the viscoelastic relaxation on the earthquakes' interaction on the KF.

By examining the combined co- and post-seismic stress change, we test whether the hypothesis of earthquake triggering is applicable to the KF. Following a scheme for classifying earthquake triggering based on the ΔCFS on rupture plane (Heidbach and Ben-Avraham, 2007), we find that among the four earthquakes succeeding the 1937 Tuosuo Lake earthquake, three events show potential triggering due to both the maximum and average ΔCFS values ≥0.01 MPa (Table 2), a threshold value for earthquake triggering. However, if we apply this scheme for the ΔCFS at the hypocenter, 2 out of 4 events are potential examples of the earthquake triggering (Table 2). For the 2001 Kokoxili earthquake, there are two proposed hypocenters: one by Harvard CMT and the other by USGS. The uncertainty of the earthquake location leads to a controversy: triggering is applicable for Harvard CMT hypocenter, while negative for USGS one.

It should be noted that, however, all the four subsequent events occurred in the regions that experienced positive stress loading, suggesting that the preceding earthquakes prompt the occurrence of the subsequent ones.

4.2 Stability of the results

We tested the stability of our results by comparing different slip models, rheological assumptions and friction coefficients of faults.

4.2.1 Slip models

The slip models of two earthquakes, 1937 Tuosuo Lake earthquake and 1997 Manyi earthquake, are highly debated. For the 1937 Tuosuo Lake earthquake two different co-seismic slip models exists (Li et al., 2006; Guo et al., 2007). The greatest controversy between two models is the western termination of the rupture, which leads to different estimations of the rupture length. Since no convincing evidence is available due to the degradation and overlap of older earthquakes, it is difficult for us to discern which one is more realistic. We applied two models to conduct a comparison study, and found that the slip model of Li et al. (2007) produced a ΔCFS increase which is about ~18 percent higher than that calculated by the model of Guo et al. (2007). This discrepancy may be resulted from the difference of the earthquake magnitudes predicted by two models (0.13). The most remarkable difference between two stress fields is the contrast of ΔCFS at the segment between the eastern extremity of the rupture of the 1963 Alake Lake earthquake and the western termination of the rupture of the 1937 Tuosuo lake earthquake proposed by Guo et al (2007). While the segment is significantly loaded (>0.02 MPa) using the Guo et al. (2007) model, it is completely within a stress shadow using the model suggested by Li et al. (2007). This contrast leads to significant discrepancy for evaluating the seismic hazard on this segment. More paleoseismological data are needed for elucidating this puzzle. However, it should be noted that, no matter what model is applied, our conclusion concerning earthquake triggering on the KF is applicable since the ΔCFS calculated by both models are of the same order in amplitude both at hypocenters and on the rupture surfaces of the subsequent events.

In contrast to the case of the 1937 Tuosuo earthquake, the two co-seismic slip models for the Manyi earthquake (Funning et al., 2007; Wang et al., 2007) produce similar ΔCFS, both in far- and near-fields. Therefore, our conclusion is rigorous no matter what slip model in used. And, our use of the slip model for present study is therefore justified.

4.2.2 Viscosity

Since viscoelastic relaxation is introduced, the viscosities of the lower crust and upper mantle are of importance for the stress calculation. In the present study, we set viscosities according to the results from studies on postseismic deformation (Ryder et al., 2007; Shao et al., 2008). Due to lack of continuous observation of postseismic deformation in the studied area, the viscosities of the crust and upper mantle are not well constrained. Therefore, we tried other choices of viscosities to test the stability of the results. Table 3 shows the results of the test experiments with various configurations of viscosities. It is shown that if the viscosity of upper mantle (η_m) is fixed to be 10^{20} Pa·s, the ΔCFS decreases with increase of the viscosity of lower crust (η_c). Similar situation is also found for the case of fixed-η_m. In this study, η_c and η_m were set to be 10^{18} Pa·s and 10^{20} Pa·s, respectively. If a lower value of η_c (10^{16}-10^{19} Pa·s) and η_m (10^{19} Pa·s) (Clark et al., 2000) are introduced to describe the weak crust and hot mantle of the Tibetan Plateau, our results of ΔCFS are substantially underestimated, and a higher value can be expected. It can be inferred that our conclusion for earthquake triggering on the KF is rigorous under most choices of the viscosities.

4.2.3 Coefficient of friction

The selection of an appropriate value for the apparent coefficient of friction μ′ is of importance for the model application, as it modulates the contribution of the normal stress to the ΔCFS. In general, μ′ is taken to be different values for different types of faults:

higher values for thrust (~0.8) and normal (~0.6) faults, while lower values (0.2-0.4) for strike-slip faults. Since the KF is a strike-slip fault with significant cumulative slip, shear stress changes, in this kind of environment, dominate over normal ones, and ΔCFS is basically governed by shear component. Therefore, we chose low value of μ′ (0.4) for stress modelling and performed tests with other values (0 and 0.6) to elucidate its impact. The numerical results show that some changes (<10%) were found in the calculated stress field. Compared with the uncertainties of the other parameters, such as rheology and slip distribution, the influence of the friction coefficient is relatively small, especially for a strike-slip fault.

4.3 Stress accumulation and seismic hazard on the KF

We extended our calculation to year 2040 to study how the ΔCFS accumulates on the KF in the coming decades (Fig. 5). The most remarkable feature of the cumulated ΔCFS on the fault is the existence of five positive ΔCFS zones, A-E (Fig. 3f and Fig. 5). Since the regions where segments A and B are located are almost unpopulated, the stress accumulation on these segments and the seismic hazard in the areas were not discussed in the present study. The calculated ΔCFS pattern on the segment D is uncertain, depending on the choice of slip models. The lack of necessary information on paleoseismology and micro-seismicity prevents us to explore more quantitative study. So, the state of ΔCFS and hazard of segment D is left as an open question. We focused our interest on the other segments, C (XDS) and E (MMS), which have not experienced any significant large earthquake over ~300 yr (Van der Woerd et al. 2002b) and ~2500 yr (SBQP, 1999; Wen et al., 2007), respectively.

Fig. 6 shows the evolution of the cumulative ΔCFS on the segment XDS and MMS, which are two major seismic gaps on the KF (SBQP, 1999; Wen et al., 2007). Although the average earthquake recurrence interval and the age of most recent earthquake (MRE) are ambiguous due to uncertainties of evidences, the XDS is believed to be of great potential of large earthquake (~Mw7.6) (Guo et al., 2006).

The evolution of the cumulative ΔCFS on XDS is displayed in Fig.6a. It is shown that, ignoring the abnormal jump due to edge effects and rupture configuration, the main part of the segment will be experienced a positive ΔCFS of ~0.25 MPa, which is much higher than the threshold of earthquake triggering. Although the ΔCFS was raised about 0.7-0.8 MPa by coseismic slip of the 1963 and 2001 earthquakes, the postseismic relaxation plays a more important role for increasing the ΔCFS. As time progresses, the postseismic relaxation is expected to be dominative for raising the seismic hazard on this segment.

It is shown that the stress accumulation on the MMS was initiated by the coseismic slip of the 1937 Tuosuo Lake earthquake, with ΔCFS of 0.15 MPa at its western extremity and about 0.02 MPa on the western part of the segment (Fig. 6b). As a result, the ΔCFS on MMS is substantially enhanced with maximum (>0.4 MPa) at its western extremity. The amplitude decreases rapidly from west to east, and tends to be a relatively small value (~0.2 MPa). In contrast to the ΔCFS on XDS, the postseismic relaxation is near completion in the following decades, although it raised ΔCFS up to 0.1 MPa in the evolution path. The steady tectonic loading will dominate the build-up process of ΔCFS. And the western extremity, Maqin, may be of the most potential of seismic hazard.

5. Discussion and Conclusions

We calculated the evolution of ΔCFS in the KF area due to five earthquakes of M≥7, by integrating coseismic and postseismic stress change together with tectonic loading. We found that all four earthquakes succeeding the 1937 Tuosuo Lake earthquake were encouraged by positive ΔCFS Two or three out of the four subsequent earthquakes were potentially triggered by preceding events, depending on the choice of classification scheme for earthquake triggering. However, we inferred that the static stress transfer is not the major control for the occurrence of subsequent events, and the postseismic viscoelastic relaxation process is more important in the stress transfer and accumulation following large earthquakes.

From the cumulative ΔCFS on the KF, we identified five segments with positive ΔCFS. Two segments, XDS and MMS, were emphasized for seismic hazard, as they has not experienced any significant large earthquake over the past several hundred years and may be loaded with ΔCFS over 0.4 MPa in the following decades.

It should be noted that our results have ambiguities due to high uncertainties of slip models, viscosities of lower crust and upper mantle, as well as incomplete catalogue of historical earthquakes. More evidences from paleoseismology, further details on rock properties, as well as fault slip rate are essential to further constrain the state of stress and to assess in particular the seismic hazard of the deteced seismic gaps.

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Figure captions

Figure 1. Location map of the Kunlun fault region and spatial-temporal migration of five M≥7.0 earthquakes along the fault during the period 1937 to 2001. Epicenter locations (grey stars), event date and focal mechanisms are summarized in Table 1. Solid lines are faults and thick solid lines are ruptured segments. Numbers inside open rectangles indicate fault slip rates. Labels inside open rectangles are names of cities and towns. Locations of cities and towns are indicated by symbols (solid dot: population 10-50 thousand; up solid triangle: 50-100 thousand; down solid triangle: 100-200 thousand; solid square: >200 thousand). Inset shows the overview of the study reagion and indicates with a black star the epicenter location of Mw7.9 Wenchuan earthquake.

Figure 2. Horizontally stratified model comprised of elastic upper crust, viscoelastic lower crust and viscoelastic mantle. V_P is the velocity of P wave. μ is the shear modulus. ρ is the rock density, and η is viscosity (η_c, crustal viscosity; η_m, mantle viscosity). η_c and η_m are set to be 1×10^{18} and $\times10^{20}$Pa·s, respectively. and other values of viscosities are used for comparison and stability tests.

Figure 3. Evolution of the Coulomb Failure Stress changes at the depth of 10 km since 1937. Thick lines are faults. Thick green, red and white lines represent the segment of the next earthquake rupture, the current earthquake rupture and the previous ruptured segments, respectively. Figures labelled from a to f are snapshots at different time: (a) immediately after the 1937 event; (b) immediately before the 1963 event; (c) immediately before the 1973 event; (d) immediately before the 1997 event; (e) immediately before the 2001 event; (f) current state of the change in Coulomb Failure Stress in year 2008.

Figure 4. Co- and combined (co- and post-seismic) change of Coulomb Failure Stress from just before 1963 Alake Lake earthquake to just before 2001 Kokoxili earthquake as a function of time for each hypocenter listed in Table 1.

Figure 5. Coulomb Failure Stress state of the KF in year 2040. Displayed are the cumulative ΔCFS calculated for the varying orientation of each fault in 1-km steps. The ΔCFS values (a) include co- and post-seismic stress changes; and (b) combined stress change (co-, post-seismic stress change and tectonic loading). Units A-E are the five segments on which ΔCFS is positive. Meanings of symbols and labels refer to Fig. 1. From west to east, the stress loaded units are labeled A) Western Manyi; B) Manyi-Kunlun Transition Zone; C) Xidatan-Dongdatan Segment (XDS); D) Alake Lake-Tuosuo Lake Segment (ATS) and E) Maqin-Maqu Segment (MMS).)

Figure 6. Evolution of the cumulative ΔCFS (co- and post-seismic) on the segments of XDS (a) and MMS (b). Line 1937 indicates the ΔCFS just after 1937 Tuosuo Lake earthquake. Lines with number of year attached by “a” represent the ΔCFS immediately before the earthquake, and those attached by “b” immediately after the earthquake. Line 2040 represents the ΔCFS state of the year of 2040.

Tables

Table 1 *Source parameters of the sequence of earthquakes used in this study*

Y/M/D	Latitude (°N)	Longitude (°E)	Strike/Dip/Rake (°)	Length (km)	Magnitude	M_0 (10^{18} Nm)	Slip[#] (m)	Ref.	Location
1937/01/07	35.40	97.69	110/70/15	150	M7.5	500	4.1	1, 2, 3	Tuosuo L.
1963/04/19	35.53	96.44	277/80/-10	40	M7.0	32	1	1, 4	Alake L.
1973/07/14	35.18	86.48	81/60/-35	66[*]	M7.3	79.2	1.5[*]	2, 4, 5	Manyi
1997/11/08	35.25	87.25	76/90/-5	170	Mw7.6	252~284	-	6, 7	Manyi
2001/11/14[a]	35.82	92.85	99/90/5	400	Mw7.8	592	-	8	Kokoxili
2001/11/14[b]	35.95	90.54							

The references used are: 1 Guo et al., 2007; 2 Molnar and Deng, 1984; 3 J. Van der Woerd, et al., 2002; 4 Molnar and Lyon-Caen, 1989; 5 Molnar and Chen, 1983; 6 Funning et al., 2007; 7 Xu, 2000; 8 Lasserre et al., 2005.

#: Slip amplitudes of 1937, 1963 and 1973 earthquakes were estimated by assuming a locking depth *w* of 20 km using the empirical relations of Wells and Coppersmith (1994)

*: The Length of the rupture and the slip amplitude were estimated by using empirical scaling laws by Wells and Coppersmith (1994).

a and b: Locations of hypocenter were given by Harvard CMT (a) and USGS (b).

- Slip distributions were given by cited references.

Table 2 Accumulated ΔCFS at hypocenters and along rupture of earthquakes on KF

No.	Date	Lat. (°N)	Long. (°E)	At hypocenter			Along rupture plane			Location
				$\Delta\sigma_c$ (MPa)	$\Delta\sigma_{c+p}$ (MPa)	$\Delta\sigma_t$ (MPa)	$\Delta\sigma_{max}$ (MPa)	$\Delta\sigma_{ave}$ (MPa)	P (%)	
1	1937-01-07	35.40	97.69	-	-	-	-	-	-	Alake L.
2	1963-04-19	35.53	96.44	0.014	0.086	0.036	0.11	0.06	100	Tuosuo L.
3	1973-07-14	35.18	86.48	4.9×10^{-5}	4.5×10^{-4}	0.051	4.52×10^{-4}	3.14×10^{-4}	0	Manyi
4	1997-11-08	35.25	87.25	0.013	0.038	0.083	0.25	0.063	98	Manyi
5	2001-11-14[a]	35.82	92.85	0.001	0.01	0.21	0.04	0.014	42.7	Kokoxili
	2001-11-14[b]	35.95	90.54	0.001	0.0046	0.069				

(a: Epicenter of the Mw7.8 Kokoxili earthquake given by the Harvard CMT; b: Epicenter given by USGS.)

P: Percentage of rupture length with $\Delta\sigma \geq 0.01$MPa;

$\Delta\sigma_c$: coseismic CFS change; $\Delta\sigma_{c+p}$: coseismic+postseismic CFS change;

$\Delta\sigma_t$: stress change due to tectonic loading; $\Delta\sigma_{max}$: Maximum of $\Delta\sigma c+p$;

$\Delta\sigma_{ave}$: Averaged $\Delta\sigma c+p$

Table 3 Accumulated ΔCFS (co- and post-seismic) at hypocenters of earthquakes on KF

Viscosiy ($\times 10^{18}$ Pa·s)		$\Delta\sigma_{c+p}$ (MPa) on hypocenter				
		1963	1973	1997	2001a	2001b
η_m=100	η_c=0.5	0.095	11×10^{-4}	0.042	0.012	0.0055
	η_c=1.0	0.086	4.5×10^{-4}	0.038	0.010	0.0046
	η_c=10	0.047	0.85×10^{-4}	0.021	0.0037	0.001
η_c=1	η_m=0.5	0.1046	2.55×10^{-3}	0.046	0.018	0.0096
	η_m=1.0	0.1016	2.1×10^{-3}	0.045	0.016	0.0084
	η_m=5.0	0.092	1.1×10^{-3}	0.041	0.012	0.0058
	η_m=10	0.089	0.78×10^{-3}	0.04	0.011	0.0052
	η_m=100	0.086	0.45×10^{-3}	0.038	0.01	0.0046